

West Antarctic Ice Sheet retreat in the Amundsen Sea driven by decadal oceanic variability

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Mass loss from the Amundsen Sea sector of the West Antarctic Ice Sheet has increased in recent decades, suggestive of sustained ocean forcing or ongoing, possibly unstable response to a past climate anomaly. Lengthening satellite records appear incompatible with either process, however, revealing both periodic hiatuses in acceleration and intermittent episodes of thinning. Here we use ocean temperature, salinity, dissolved-oxygen and current measurements taken from 2000-2016 near Dotson Ice Shelf to determine temporal changes in net basal melting. A decadal cycle dominates the ocean record, with melt changing by a factor of ~4 between cool and warm extremes via a non-linear relationship with ocean temperature. A warm phase that peaked around 2009 coincided with ice shelf thinning and retreat of the grounding line, which re-advanced during a post-2011 cool phase. Those observations demonstrate how discontinuous ice retreat is linked with ocean variability, and that the strength and timing of decadal extremes is more influential than changes in the longer-term mean state. The non-linear response of melting to temperature change heightens the sensitivity of Amundsen Sea ice shelves to such variability, possibly explaining the vulnerability of the ice sheet in that sector, where subsurface ocean temperatures are relatively high.

The West Antarctic Ice Sheet (WAIS) has recently been contributing $\sim 0.3 \text{ mm yr}^{-1}$ to global sea level rise¹, with the potential to discharge ice more rapidly into the ocean where its bed slopes downward inland to as much as 2.5 km below sea level at its centre². That configuration can be inherently unstable³ unless floating ice shelves at its margin provide sufficient restraint on ice discharge across the grounding line^{4,5}. Since buttressing is generated where ice shelves contact lateral margins and seabed shoals, ice shelf thinning reduces buttressing, accelerating the outflow of grounded ice⁶, a process that makes the WAIS highly sensitive to ocean conditions at its margin. Understanding the physical links between oceanic change and ice loss is thus critical to assessing the future of WAIS, and the potential reversibility of recent changes.

Satellite observations provide a consistent picture of mass loss from West Antarctica's Amundsen Sea sector over recent decades^{1,7,8,9}. Flow acceleration of outlet glaciers⁷ has been accompanied by inland thinning^{1,9} and retreat of grounding lines⁸, while lengthening records have revealed elevated rates of mass loss in recent years^{1,7}. Key outlet glaciers are buttressed only by small ice shelves¹⁰ that are furthermore exposed to relatively warm seawater¹¹. The thinning signature propagates inland from grounding lines⁹, with downstream ice shelves thinning faster¹², pointing to changes in buttressing triggered by ocean-driven melting as the cause.

Most studies have attributed the mass loss either to unstable retreat of the grounding lines^{13,14}, possibly triggered by a climate anomaly in the 1940s (15), or to sustained ocean forcing resulting from a past increase in the quantity of warm water on the continental shelf¹⁶ or a long-term warming of those waters¹⁷. Recent observations^{7,8,9} are, however, irreconcilable with such hypotheses. Alternating phases of rapid acceleration and steady or even decelerating flow are mostly coherent across different glaciers⁷. If all were retreating unstably or under sustained forcing, variations in the rate of change would be dictated by the geometry and flow regime of individual glaciers, and incoherent. Furthermore, current episodes of rapid thinning commenced on Pine Island and Thwaites glaciers less than 30 years ago⁹, excluding a simple link with events in the 1940s.

Oceanographic records remain short and patchy on the Amundsen Sea continental shelf, where the first full-depth profiles of seawater properties in the early 1990s revealed warm derivatives of Circumpolar Deep Water in deep, glacially-scoured troughs and a consequent high melt rate beneath Pine Island Ice Shelf¹¹. Subsequent summer cruises indicated that Pine Island melt rates tracked ocean heat content, increasing between 1994 and 2009 (18) and decreasing between 2010 and 2012 (19). The cool phase persisted for several years²⁰, causing deceleration of the glacier that has typically been interpreted as a minor perturbation to the long-term retreat^{21,22}. Getz Ice Shelf experienced a more pronounced change in melting between cool ocean conditions in 2000 and warm in 2007 (23), but the inland ice response remains undocumented. For both ice shelves, ship-based measurements have been limited by persistent sea ice fields, but near-repeat stations throughout the Amundsen continental shelf indicate spatially coherent cool and warm periods²⁴.

Melt rates inferred from observation at Dotson Ice Front

In this region, the most reliable area of summer open water is the Amundsen Sea Polynya, which forms to the north of Dotson and eastern Getz ice shelves. Ships have thus been able to access that area more often, yielding eight near-repeat sets of seawater properties close to Dotson Ice Front from 2000-2016 (Figure 1). Combining those observations shows how Dotson basal melting has evolved in response to changing ocean conditions.

The ice front water column comprises two relatively uniform layers separated by a region with higher vertical thermohaline gradients (Figure 2). The warmer and saltier bottom layer (typically near 0.5°C and 34.55), is derived from modified Circumpolar Deep Water (mCDW) that flows onto the shelf near 118°W and circulates cyclonically within the Dotson-Getz Trough^{25,26}. The colder, fresher upper layer (typically near -1.25°C and 33.8) is derived from Winter Water (WW), formed as a result of cooling, brine drainage and convection beneath growing sea ice, and having properties that vary mainly through the degree of summer warming and the depth of winter mixing. The bottom layer stands out for being coldest in 2000, but more striking temperature changes occur near

the mid-depth transition region between the two layers (Figure 2). When that thermocline is shallow (deep), significantly more (less) ocean heat is available to melt ice within the sub-Dotson cavity. Deep thermoclines in 2000 and 2012-2016 bracket shallower levels in 2006-2011, changes that are coherent with less complete records elsewhere^{18,19,23}, albeit with systematic spatial differences²⁴.

After calculating seawater density from temperature and salinity, those parameters are used to determine the circulation and transport of meltwater across the ice front (Methods). In years when ocean current profiles were directly measured, their combination with seawater properties yields independent estimates of circulation and melting. A persistent east side inflow that is warmer and deeper than west side outflow is consistent with the addition of meltwater from the ice shelf base driving a geostrophic circulation (Figure 3). More variable transports appear elsewhere across the section.

Estimates of the net meltwater flux across the sections closely track variations in mean temperature above freezing associated with changes in thermocline depth (Figure 4a). Both quantities are estimated using observations over the entire section, but exclude a variable-depth surface layer where air-sea interaction influences the water properties (Methods). Meltwater fluxes were ~4 times higher at the peak warm phase in 2009 than during cool phases before and after, when values were similar despite the change in deep temperature maximum (Figure 2a). Intermediate melt rates in 2006 and 2007 are consistent with a glaciological estimate²⁷ derived from 2003-2008 satellite data (Figure 4a). Satellite-based estimates of spatial²⁸ and temporal^{29,30} variability in Dotson Ice Shelf melting combine ice thicknesses derived from surface elevation data, a procedure that magnifies uncertainties by an order of magnitude, with surface accumulation and firn compaction rates derived from models that lack independent verification²⁹ or assumed constant³⁰. Although subject to their own methodological uncertainty, our estimates are independent of regional climate models and assumptions about firn density.

Ice sheet changes driven by melt rate variability

The 16-year history of melting explains the recently-observed behaviour of Dotson Ice Shelf and its tributary ice streams, especially Kohler Glacier (Figure 4b). Surface accumulation and iceberg calving make small, mutually-cancelling contributions to the Dotson mass budget²⁷, so when melt equals ice flux across the grounding line the ice shelf is close to equilibrium. The low, near-equilibrium melt rate in 2000 was followed by 4 years of steady ice influx⁷. A period of rapid acceleration⁷, thinning³¹, and grounding line retreat⁸ then occurred from 2004-2011, when melt rates significantly exceeded equilibrium values. A cooler interval beginning in 2012 caused a steadying of the flow⁷ and re-advance of the Kohler Glacier grounding line⁸, as melting dropped below the now-elevated equilibrium value.

Such behaviour provides insight into the physical processes that link ocean forcing with ice sheet response. Thermocline shoaling induces thinning of the ice shelf, reducing its contact with margins and seabed, and weakening buttressing at the grounding line⁶, where flow accelerates in response⁷. Consequent thinning of inland ice causes retreat of the grounding line⁸, further reducing local resistance to flow³², and steepens the slope of the glacier surface, increasing local forcing of the flow³³. Those secondary effects induce further acceleration, and develop over time at a rate that is determined by the glacier bed and margin geometries, and the local resistance they provide to the flow⁹. Subsequent deepening of the thermocline reduces melting, causing ice shelf thickening and increased buttressing at the new grounding line. Inland response is now complicated by a continuing adjustment of ice thickness and flow to the earlier perturbation.

The non-linear relationship between meltwater flux and mean seawater temperature (Figure 4c) confirms theoretical inferences wherein turbulent ice-ocean heat transfer depends on both the temperature difference across the boundary layer and the shear generated by the far-field geostrophic current, itself driven by temperature-related density differences (Methods). A similar relationship exists in numerical models of sub-ice circulation applied to idealised ice and seabed

geometries³⁴. That non-linearity has important implications for ice shelf vulnerability to changing ocean conditions, enhancing melt rate sensitivity as the mean state temperature rises. For a given temperature increase, ice shelves will thin more rapidly in the warm SE Pacific sector than elsewhere on the Antarctic continental shelf, where subsurface seawater temperatures are typically closer to the surface freezing point.

Ocean variability as a driver of change in the Amundsen Sector

Models and observations give a consistent picture of winds over the continental shelf as the main driver of thermocline depth variability and consequent changes in melting throughout the eastern Amundsen Sea^{22,35,36}. Various processes including wind-forced variability of mCDW inflow³⁷ and near-coastal densification^{20,38} and downwelling of WW³⁹ have been invoked. If critical wind changes are associated with regional circulation anomalies triggered from the tropical Pacific^{29,40}, forcing during prior decades would have been characterised by similar cycles²². Coupled with the non-linear response of ice shelf melting, such pronounced variability makes the Amundsen Sea sector of the WAIS particularly susceptible to decadal changes in ice shelf buttressing.

The region-wide observations of intermittent acceleration⁷ and more sustained, but episodic thinning⁹ are explicable in terms of such forcing (Figure 5). Warm phases are characterised by rapid acceleration in ice flow, as ice shelves thin and buttressing decreases. Flow steadies or decelerates during cool phases, with exceptions because individual glacier bed geometry determines the size of the evolving inland response to the preceding acceleration. The resilience of Kohler Glacier through the recent period of elevated melting beneath Dotson Ice Shelf indicates that a grounding line pinned on a prominent seabed rise⁸ can limit the response to reduced buttressing and allow recovery during a subsequent cool phase. With a less stable grounding line, acceleration resulting from reduced buttressing can initiate a short period of unstable retreat, triggering an inland-propagating wave of thinning⁹, and further acceleration that may over-ride a cooling-induced increase in ice shelf buttressing. Evidence (Figure 5) suggests that episodes of retreat occurred in

the 1940s, the 1970s and the 1990s on Pine Island Glacier^{9,15}, in the early 2000s on Thwaites and Haynes glaciers⁹, and with less certain timing on the glaciers feeding Dotson and Crosson ice shelves⁹.

Longer-term change may underlie the pronounced decadal variability reported here. That might account for the mean ocean state over the 2000-2016 period being warm enough to induce thinning of Dotson Ice Shelf (Figure 4b). However, inland thinning of Kohler Glacier⁹ indicates a retreat episode prior to 2000 (Figure 5). Associated deepening of the grounding line would have exposed more of the ice shelf base to the warmer mCDW, elevating the melt rate for equivalent ocean forcing. The absence of significant inland thinning on Pine Island Glacier prior to the 1990s retreat⁹ suggests that a near-equilibrium state was re-established there following major perturbations in the 1940s and 1970s (15).

It now appears that recent accelerated mass loss from the Amundsen Sea sector of the WAIS has not resulted from progressive ocean warming or unstable ice retreat, but rather from a combination of analogous processes whereby successive warm intervals trigger episodic retreats of the most vulnerable grounding lines, adding to a longer-term inland response. Over recent decades, pauses in acceleration and minor decelerations in Amundsen Sea outlet glaciers during cool periods (Figure 5) may have reduced the rate of ice loss from West Antarctica, as documented since 2011 for the current cool phase⁴¹. Determining the magnitude of committed loss from past and future warm episodes will require better understanding of ice shelf and outlet glacier response to, and the origins of, the strong temporal variability of the mid-depth thermocline that dominates the recent record of shelf water properties.

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268 **Author contributions**

269 AJ, DS, PD and SJ conceived the study. DS, SJ, TWK, SHL, HKH and SS planned and led the data
270 collection. DS, PD, and TWK processed the data. AJ, DS and PD undertook the data analyses and
271 derivation of the final results. AJ prepared the manuscript. All authors discussed the results and
272 implications and commented on the manuscript at all stages.

273 **Competing financial interests**

274 The authors declare no competing financial interests.

275 **Materials and correspondence**

276 Correspondence and material requests should be addressed to A. Jenkins (ajen@bas.ac.uk), or to the
277 contacts listed in Supplementary Table 1 for more specific cruise information.

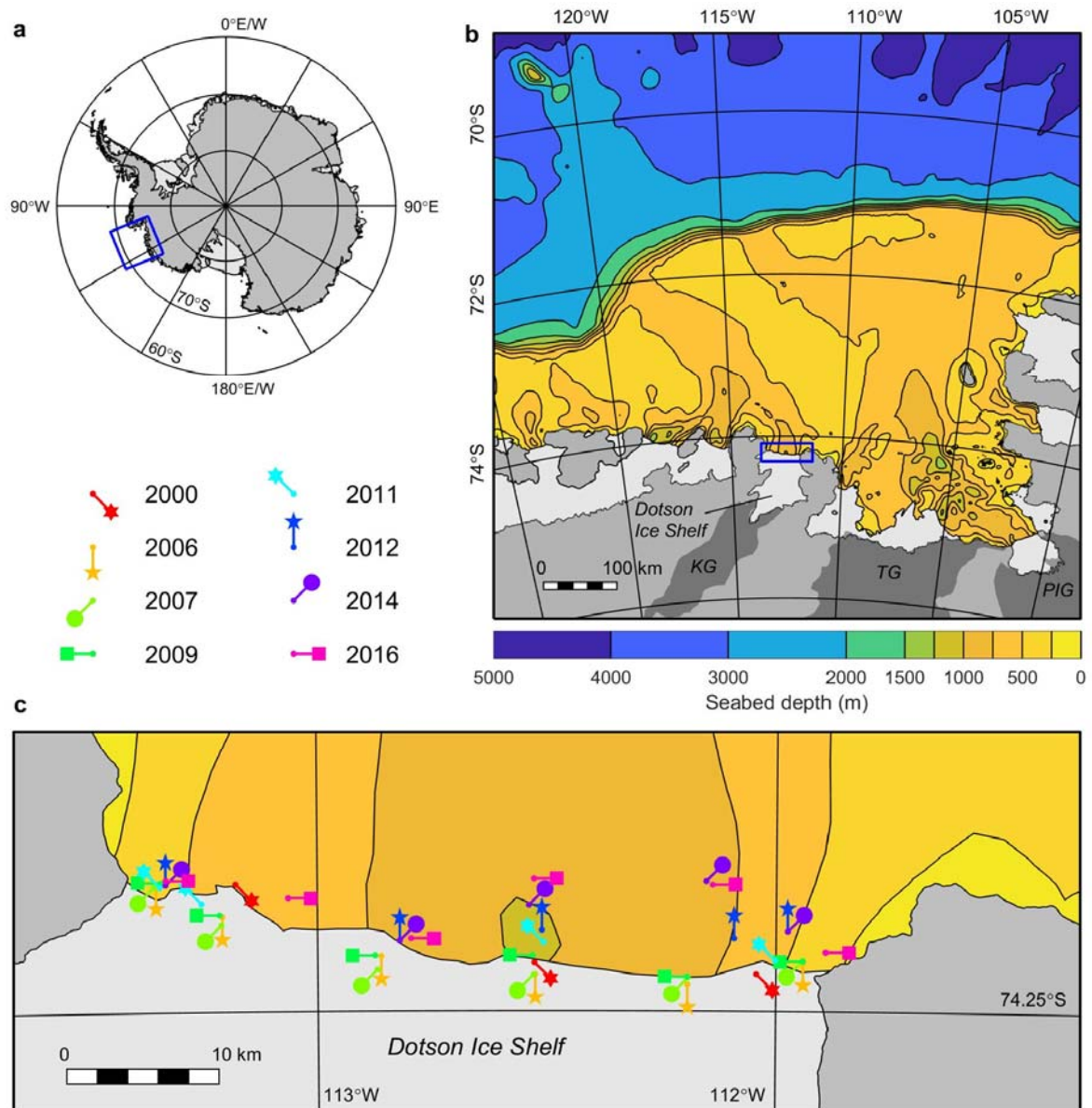


Figure 1 | Locations of Amundsen Sea observations used in this study. (a) Map of Antarctica indicating grounded ice sheet (darker shading), floating ice shelves (lighter shading) and study area (blue box). (b) Enlargement of study area showing regional bathymetry² and catchments (darker shading) of Kohler, Thwaites and Pine Island glaciers (KG, TG, PIG). (c) Enlargement of area near the moving Dotson Ice Front (blue box in panel b) showing locations and years of summer (Dec-Mar) vertical profiles of seawater properties.

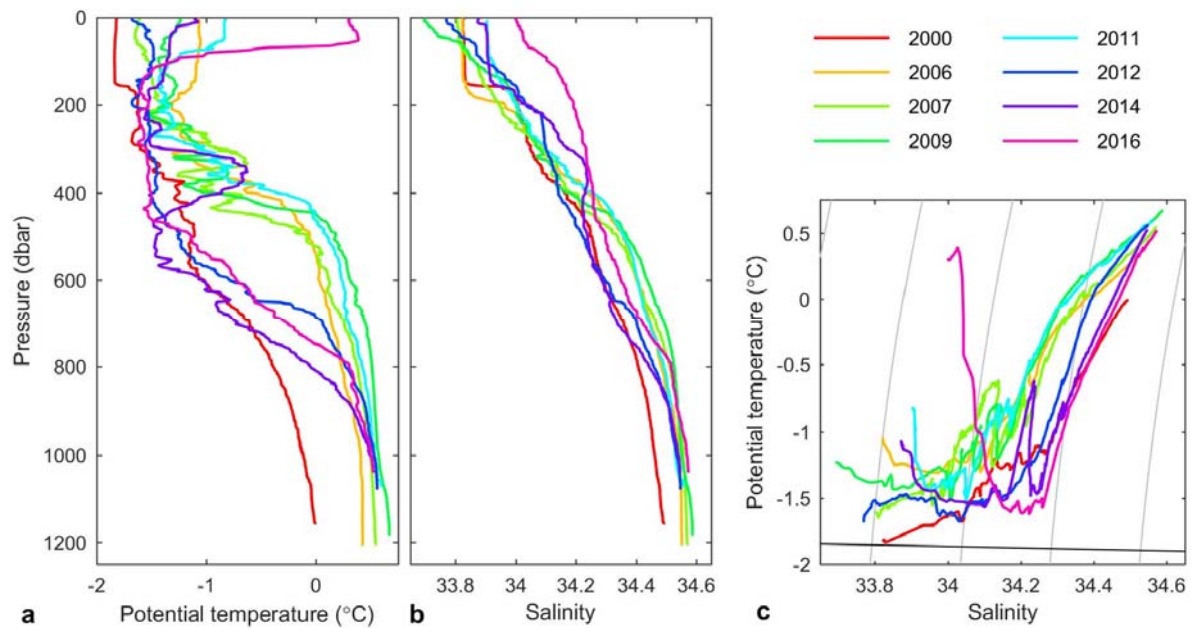


Figure 2 | Potential temperature and salinity at Dotson Ice Front. (a) Vertical profiles of potential temperature recorded at the deepest station sampled in each summer; near-repeats, centrally located in the trough near 112.5°W (Figure 1c). (b) Vertical profiles of salinity recorded at the same stations. (c) Potential temperature versus salinity for the stations in panels a and b. Grey lines connect points of equal potential density (1027 to 1027.8 kg m⁻³ in 0.2 kg m⁻³ increments), while the black line indicates the surface freezing point.

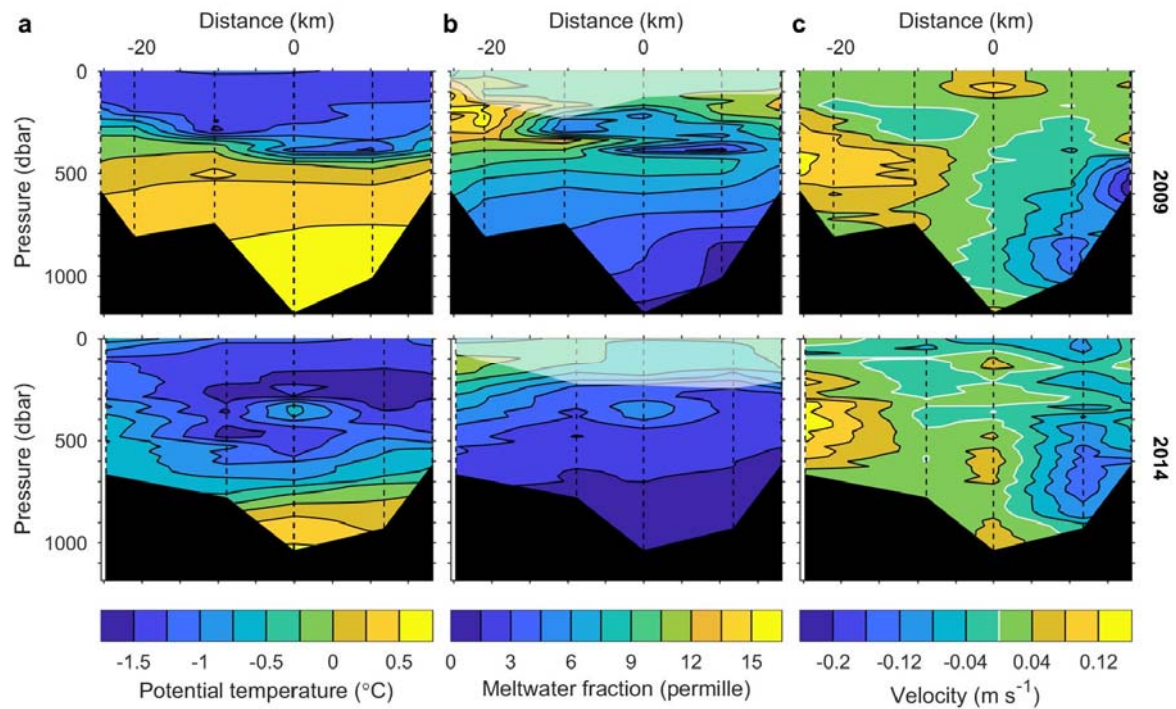


Figure 3 | Cross-sections of potential temperature, meltwater fraction and current speed perpendicular to the ice front in contrasting years. (a) Potential temperature at stations (vertical dashed lines) in 2009 (top) and 2014 (bottom). Distance relative to the central trough station (Figure 2) is positive eastward, and black shading shows the seabed, linearly interpolated between stations. (b) Meltwater fraction derived from potential temperature, salinity and dissolved oxygen concentration (Methods), lighter where the calculation is unreliable due to the influence of air-sea interactions. (c) Lowered Acoustic Doppler Current Profiler speeds perpendicular to the ice front, positive for northward outflows.

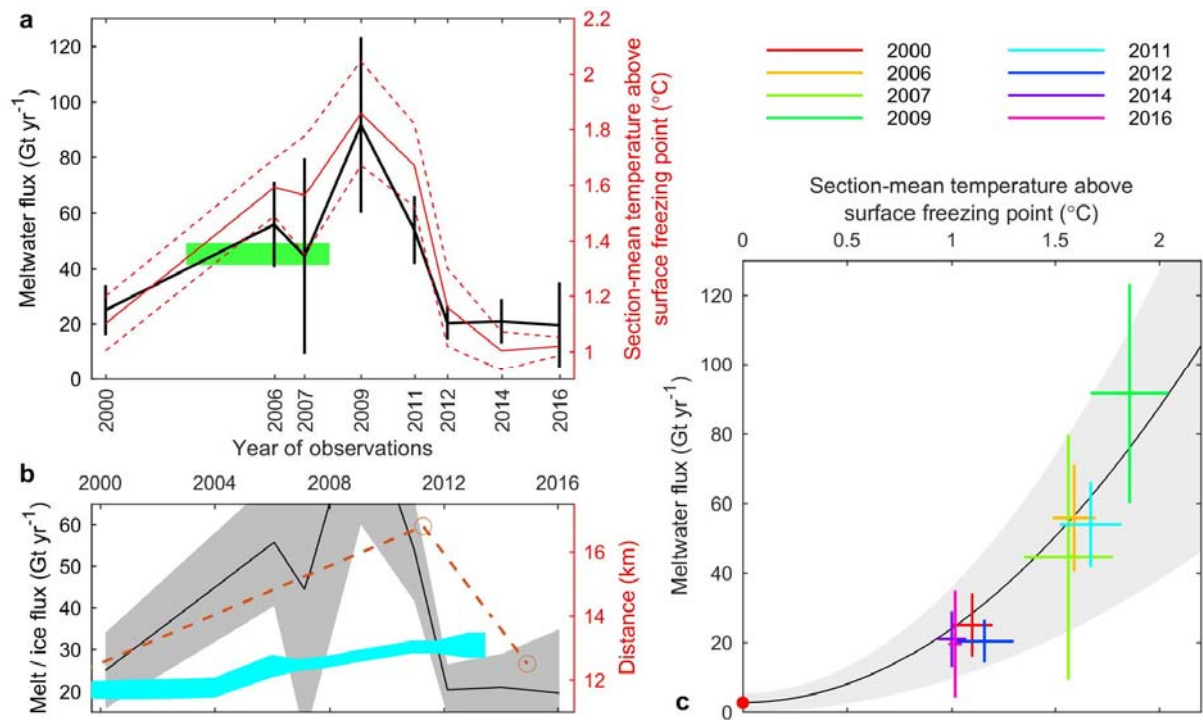


Figure 4 | Meltwater flux and mean ocean temperature at Dotson Ice Front. (a) Time series of meltwater flux (black, one standard deviation bars) and mean temperature above the surface freezing point (red, one standard deviation dashed lines), with melt flux derived from 2003-2008 satellite data²⁷ (green). (b) Meltwater flux (black), mass flux across Dotson Ice Shelf grounding line⁷ (cyan) and relative position of Kohler Glacier grounding line⁸ (red). (c) Meltwater flux versus mean temperature above the surface freezing point (colour-coded by year). The black line is a theoretical quadratic relationship (Methods) with uncertainties (grey shading), using an additional point (red dot) estimated from observations in cold water sectors of Antarctica²⁷.

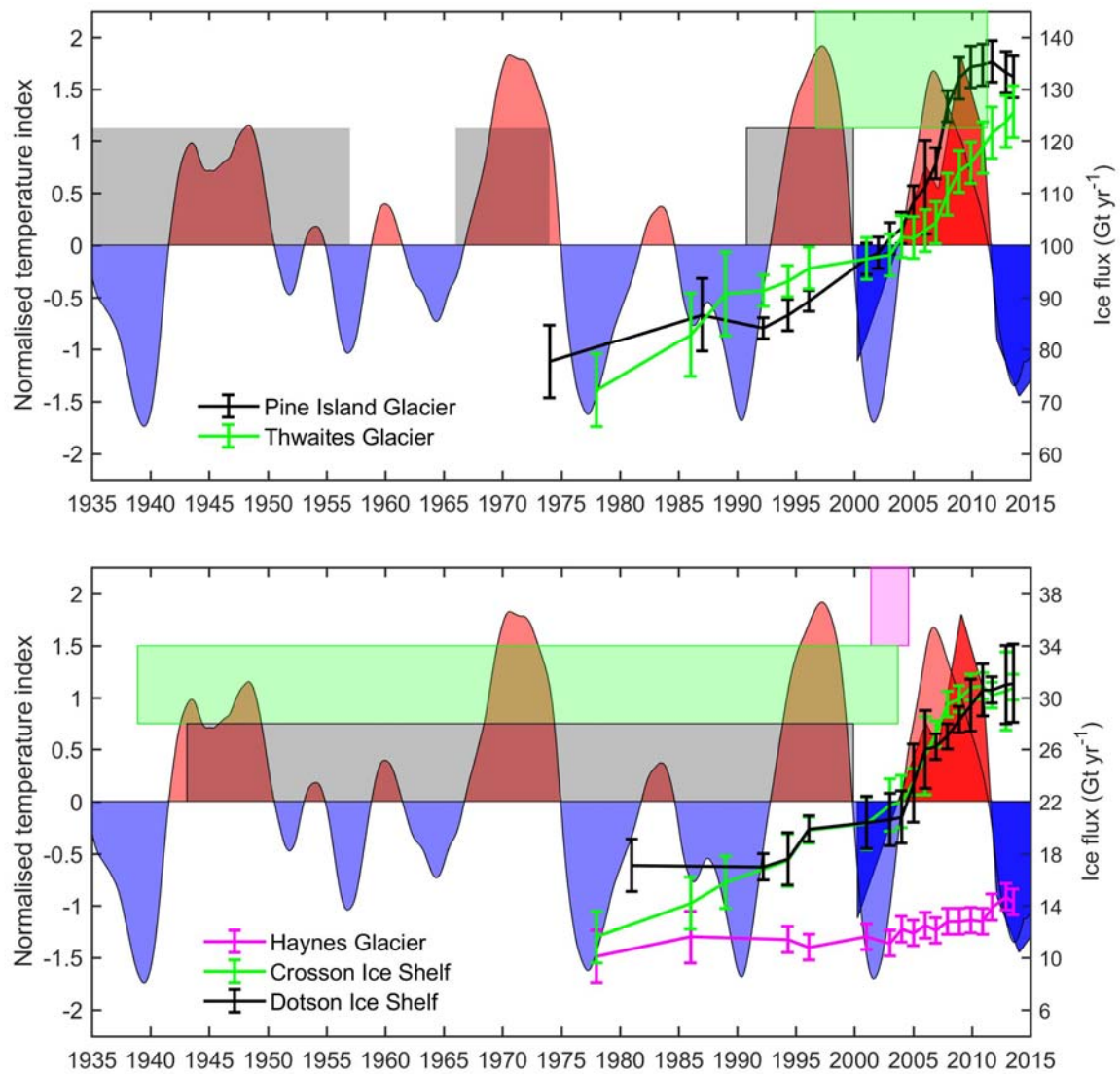


Figure 5 | Multi-decadal history of ocean forcing and outlet glacier response in the eastern Amundsen Sea. Time series of glacier outflow changes⁷ (right-hand axis) and ocean forcing (red = warm conditions; blue = cool conditions) as documented here (darker shading) and inferred (lighter shading) from central tropical Pacific sea surface temperatures²² (left-hand axis, both normalised). Shaded boxes (outlined and colour-coded by glacier) indicate the range of estimated times for the initiation of the most recent phase of rapid thinning at the grounding lines⁹, while boxes without outlines are inferred times of initial and final detachment of Pine Island Glacier from a submarine ridge¹⁵.

Methods

Data collection

Data were collected from three marine research vessels during eight summer visits to the Amundsen Sea (Supplementary Table 1). All cruises used Sea-Bird Scientific SBE 911plus Conductivity-Temperature-Depth (CTD) profiling systems, with consistent recording and processing procedures. On seven cruises, dissolved oxygen (DO) sensors were added to the system, and on five, Lowered Acoustic Doppler Current Profilers (LADCP) were mounted on the CTD frame. Scalar data were averaged into 1- or 2-dbar pressure bins, while current data were averaged into larger bins of typically 20 dbar. Most section profiles were regularly spaced within a few hundred metres of the Dotson Ice Front (Figure 1 and Supplementary Figure 1).

Meltwater fraction calculations

As ice melts into seawater the resulting water mass becomes cooler, fresher and richer in dissolved oxygen than the original. Most of the cooling occurs because the seawater supplies the latent heat required for the phase change, freshening because of the addition of meltwater, and dissolved oxygen concentrations rise as air bubbles trapped in the ice go into solution. Any combination of potential temperature, salinity and dissolved oxygen can be used to diagnose the meltwater content as the third component of a mixture incorporating modified Circumpolar Deep Water (mCDW) and Winter Water (WW) as the other end members⁴² (Supplementary Table 2). Ice properties are assumed to be constant, whereas mCDW and WW properties are cruise specific and are defined by reference to the distribution of data in property-property plots (Supplementary Figure 2).

In property-property space, the data lie within a triangle defined by straight lines originating from the mCDW properties⁴². One line connects the mCDW and ice end points and indicates seawater properties that arise from ice melting into “pure” mCDW. A second line connecting mCDW and WW end points represents properties that arise from mixing between the “pure” WW and mCDW.

Mixtures on the second line contain no meltwater and define the “ambient” water column that interacts with the ice. The distance that a point lies from the “ambient” line indicates the meltwater content that can be quantified as⁴²:

$$\phi = \frac{(\chi^2 - \chi_{mCDW}^2) - (\chi^1 - \chi_{mCDW}^1)(\chi_{WW}^2 - \chi_{mCDW}^2)/(\chi_{WW}^1 - \chi_{mCDW}^1)}{(\chi_{ICE}^2 - \chi_{mCDW}^2) - (\chi_{ICE}^1 - \chi_{mCDW}^1)(\chi_{WW}^2 - \chi_{mCDW}^2)/(\chi_{WW}^1 - \chi_{mCDW}^1)}$$

where ϕ is the meltwater fraction, χ^i represents any of potential temperature, salinity or dissolved oxygen, and subscripts indicate defined water mass properties (Supplementary Table 2).

WW is formed when heat loss to the atmosphere causes freezing at the sea surface. As the sea ice grows, salt is rejected from the solid phase, increasing surface water salinity and driving convection. The result is a deep mixed layer with a temperature at the surface freezing point and a salinity that varies from year to year, depending on the amount and salinity of sea ice grown during the winter. Formed by air-sea interaction at the surface, WW is rich in dissolved oxygen, but not necessarily saturated as the consolidating ice cover presents a barrier to air-sea gas exchange. Since “pure” WW, having a temperature at the surface freezing point, is absent from the summertime ice front sections, we estimate a range of possible properties (Supplementary Table 2, Supplementary Figure 2) and investigate the sensitivity of results to that property range. The mCDW properties are defined as the intersection between the ice/mCDW mixing line, which connects the ice properties with one edge of the data “triangle”, and the WW/mCDW mixing line, which connects the mid-range of the WW properties with the other edge of the data “triangle” (Supplementary Figure 2).

In some of the cold years we find that the mCDW properties thus defined do not correspond to the warmest waters near the ice front, indicating a lack of direct interaction with the ice. This points to recirculation deeper than the ice shelf draught, or exclusion from the inner cavity by a seabed shoal⁴³, as on Pine Island Glacier^{44,45}. Heat from those waters could still contribute to melting if mixed into the thermocline. In other years the mCDW properties are warmer and saltier than any waters near the ice front, an indication that some seawater arriving at the ice front has formed from

ice melting into the warmer and saltier mCDW in Pine Island Bay²⁴. Those waters flow westward along the continental shelf^{45,46} and carry meltwater beneath Dotson Ice Shelf.

Profiles of meltwater fraction, calculated using mid-range WW properties, are shown for all stations in Supplementary Figure 3. Since interaction with the atmosphere can influence all three tracers, the assumption of a three water mass mixture breaks down near the sea surface. In principle a fourth, surface water mass could be defined, but in practice the surface properties are too spatially heterogeneous to be defined as a single water mass. Instead we use the scatter of results derived from the three separate calculations of meltwater fraction to indicate where atmospheric interaction (or another process) makes the results unreliable (Supplementary Figure 3). This exclusion will lower net melting estimates if cavity outflows are light enough to upwell into the surface layer. Sections of potential temperature, salinity and dissolved oxygen are shown in Supplementary Figure 1 and derived meltwater fractions are shown in Supplementary Figure 3.

Meltwater transport calculations

The net flux of meltwater away from the ice shelf is calculated by combining estimates of meltwater fraction and currents perpendicular to the ice front. While the latter can be observed directly, as during some cruises (Supplementary Table 1), inherently large measurement uncertainties are compounded by high-frequency variability associated with tides, waves and eddies that contaminate the longer-term mean flow relevant to the calculation of meltwater transport. However, higher-frequency processes have a smaller impact on the density structure derived from measurements of temperature and salinity, of greater value in estimating the longer-term mean circulation⁴⁷.

At spatial scales larger than the local internal Rossby radius (a few kilometres) the pressure gradient that drives the flow is balanced primarily by the Coriolis force it generates, allowing across-section currents to be estimated from along-section pressure gradients. To make direct use of the pressure recorded by the CTD, it is convenient to work in an isobaric framework, in which the geostrophic currents can be derived from the slope of isobars relative to geopotential surfaces. As isobar depth

is the integral of specific volume (reciprocal of density) from the surface down to the relevant pressure, relative isobar slope between any two stations can readily be derived from the CTD data. Absolute isobar slope remains unknown because the slope of the sea surface relative to the geoid cannot be precisely measured, introducing a pressure-independent unknown into the calculated isobaric slopes. Calculated currents thus include a CTD-derived, pressure-dependent component, and a pressure-independent component that must be estimated in some other way:

$$v(P) = v(P_{ref}) + \frac{1}{f} \frac{\partial}{\partial x} \left(\int_P^{P_{ref}} \alpha dP \right) \Big|_P$$

where x is the along-section coordinate, v the across-section current, P is pressure, α is specific volume, f the Coriolis parameter, and ref refers to an arbitrary reference pressure where the flow associated with the absolute slope of the isobar must be determined⁴⁷.

The conventional solution is then to select a pressure level where it is physically reasonable to assume the isobar is parallel to the geoid and hence the velocity is zero. For the Dotson Ice Front sections, such a level of zero motion might be the maximum depth on each station at which meltwater is found (Supplementary Figure 3). Zero motion at that level would be consistent with a pure overturning circulation with warm inflow at depth and outflows of glacially-modified water at shallower levels. Given the large scale of the ocean cavity beneath Dotson Ice Shelf compared with the internal Rossby radius, we should anticipate a significant horizontal, geostrophic circulation accompanying the overturning^{18,19}. The magnitude of the horizontal circulation can be estimated by adapting a procedure for inferring reference level currents⁴⁸. Since the CTD sections completely enclose a volume of the ocean, integral constraints on the circulation across the section can be derived from the conservation of mass and scalar properties, along with a steady state assumption within that volume. Flow between individual station pairs can then be adjusted by adding to each a non-zero $v(P_{ref})$ that makes the overall circulation compatible with the constraints. With typically fewer constraints than unknown reference velocities, yielding infinitely many solutions that satisfy

the constraints exactly, the adopted strategy is to choose the solution that minimises the sum of the squares of the reference velocities (i.e. that is closest, in terms of the L_2 norm, to the original circulation with zero reference velocities).

The next problem is that the conventional approach of imposing zero mass, heat and salt fluxes at the ice front constrains the solution to give a net meltwater flux of zero across the section, in turn requiring a novel strategy to overcome that limitation⁴⁷. Knowing the ice properties (Supplementary Table 2), the impact of meltwater addition on all conservation equations (mass, heat, salt and dissolved oxygen, where measured) can be quantified in terms of the yet-to-be-determined meltwater flux. The mass conservation equation can then be used to eliminate the unknown melt rate from the other equations. The system yields as many constraints as there are tracers observed, effectively balancing the budgets associated with the circulation of meltwater-free source waters, and the resulting imbalance between total inflow and outflow is the sought-after meltwater flux (Supplementary Table 3). Another estimate of the meltwater flux can be obtained from the net transport of meltwater fraction, quantified as described above, associated with the derived circulation. Both estimates yield the net meltwater flux across the section, with the latter additionally giving the meltwater concentration in the inflow that is sourced from elsewhere (Supplementary Table 3) and the total in the outflow. The two estimates of the melt added from the ice shelf are independent, since the first requires only that integrated inflow and integrated outflow properties be connected by a meltwater mixing line. It is thus formally independent of the choice of WW and mCDW properties, although indirect use is made of the scatter in meltwater fraction calculations to determine how much of the water column to exclude from the budget calculations. The scatter indicates where air-sea exchange exerts an influence on water properties (Supplementary Figure 3), and that part of the upper water column is excluded from budget calculations that assume the only changes between inflow and outflow result from meltwater addition.

The initial circulation, adjustments and final circulation derived using mid-range WW properties are shown for each section in Supplementary Figure 4, with analogous results from available directly-observed currents in Supplementary Figure 5. As the largest currents associated with the undesired high-frequency variability are likely to be approximately pressure-independent, an identical technique preserves the observed depth-dependent current structure, but closes budgets by adding depth-independent velocity adjustments calculated as described above. Prior de-tiding of the observed currents has a negligible impact on the results, because tidal currents near Dotson Ice Shelf are typically much smaller than the applied adjustments.

Since some underlying assumptions, particularly that of steady, geostrophic flow, will hold only approximately, any one solution will generally provide an unreliable estimate of the circulation and ice shelf melt rate. Indeed, applying all constraints can sometimes yield excessively noisy solutions, since we are forcing inherently noisy observations to fit our assumptions exactly. We therefore generate the inverse of the constraint matrix using a Singular Value Decomposition, and we use a truncated rather than a full-rank solution if the latter is excessively noisy⁴⁸. With at most three constraints, this amounts to a choice between three (full-rank) or two (truncated) singular vectors. The procedure is repeated for all realisations of the WW properties (Supplementary Table 2), and each solution scored based on three measures of solution credibility:

$$\text{Score} = \frac{1}{3} \left\{ \left(1 - \min \left[1, \frac{v_{rms}^{adj}}{V_{rms}^{adj}} \right] \right) + \left(1 - \min \left[1, \frac{v_{max}^{adj}}{V_{max}^{adj}} \right] \right) + \left(1 - \min \left[1, \frac{\delta m}{\Delta M} \right] \right) \right\}$$

where the three criteria are the root-mean-square velocity adjustment, v_{rms}^{adj} ; the maximum absolute adjustment, v_{max}^{adj} ; the difference between the transport imbalance and the net flux of meltwater fraction across the section, δm . In each case we specify a tolerance beyond which the solution scores 0 for that criterion:

$$V_{rms}^{adj} = 0.1 \text{ m s}^{-1}; \quad V_{max}^{adj} = 0.2 \text{ m s}^{-1}; \quad \Delta M = 20 \text{ Gt yr}^{-1}$$

with levels chosen to remove the influence of the majority of physically unreasonable solutions. Additionally, the score for any solution with a meltwater flux less than zero or a circulation greater than 2 Sv is set to zero regardless of the other criteria. All solutions are shown in Supplementary Figure 6, along with the score-weighted mean and score-weighted standard deviation that provide the final results plotted in Figure 4. Further results are summarised in Supplementary Table 3.

Theoretical relationship between meltwater flux and ocean temperature

Melting at the base of an ice shelf cools and freshens the water near the ice base and creates a buoyant current that flows along the ice-ocean interface. The melt rate (m) is determined by the heat flux across the turbulent boundary layer created by current shear against the ice base, and can be quantified as⁴⁹:

$$m = \left\{ \frac{C_d^{1/2} \Gamma_{TS}}{(L_i - c_i T_{*i})/c} \right\} U T_*$$

where U is current speed, $T_* = T - T_f$, is temperature relative to the salinity-dependent freezing point at the depth of the ice shelf base, sometimes referred to as thermal driving, c is specific heat capacity, L latent heat of fusion, C_d drag coefficient, Γ_{TS} a heat exchange coefficient, and the subscript i indicates ice properties. If the drag and thermal exchange coefficients are assumed to be constant, the term in braces, which will be denoted M_0 , is approximately constant, as it depends only weakly on the ice temperature relative to the seawater freezing point.

Assuming the large-scale circulation to be in geostrophic balance, the speed of the buoyant boundary current relative to the lower layer of inflowing water, $\Delta U = U - U_{in}$, can be written⁵⁰:

$$\Delta U = \frac{g}{f} \sin \theta \Delta \rho$$

where g is gravity, f the Coriolis parameter, θ the slope of the ice shelf base relative to the horizontal and $\Delta \rho = (\rho - \rho_{in})/\rho_{in}$, the dimensionless density deficit in the boundary current. The transports in the lower, inflowing layer and the buoyant boundary current are approximately equal,

differing only because of the addition of meltwater to the latter, so if both layers are of similar thickness, the speed of the buoyant current can be approximated as:

$$U \approx \frac{\Delta U}{2}$$

The properties of the boundary current must lie somewhere along the mixing line connecting the properties of the inflow with those of the ice (the green lines in Supplementary Figure 2), so the temperature and salinity differences are related by⁴²:

$$\frac{(T - T_{in})}{(S - S_{in})} = \frac{T_{*in} + (L_i - c_i T_{*i})/c}{S_{in}}$$

Combining this expression with a linear equation of state:

$$\Delta\rho = \beta_S(S - S_{in}) - \beta_T(T - T_{in})$$

and a linear equation for the freezing point at the depth of the ice shelf base as a function of salinity:

$$(T_* - T_{*in}) = (T - T_{in}) - \lambda_1(S - S_{in})$$

leads to an expression for the density deficit in terms of the thermal driving deficit⁵⁰:

$$\Delta\rho = (T_* - T_{*in}) \left\{ \frac{\beta_S S_{in} - \beta_T [T_{*in} + (L_i - c_i T_{*i})/c]}{T_{*in} + (L_i - c_i T_{*i})/c - \lambda_1 S_{in}} \right\}$$

Once again the term in braces, which will be denoted P_0 , is approximately constant because of the weak dependence on the ice and inflow temperatures and the small range in inflow salinity.

Thermal driving in the boundary current must be less than that in the inflow, but greater than zero (if the ice shelf base is melting), a condition that can be expressed as:

$$T_* = \varepsilon T_{*in} \quad 0 < \varepsilon < 1$$

and combining all the above results leads to an expression for the melt rate that is a quadratic function of the inflow temperature:

$$m = M_0 \frac{g}{2f} \sin \theta P_0 \varepsilon (\varepsilon - 1) T_{*in}^2$$

The term ε can be related to the relative efficiency of mixing across the thermocline that separates the boundary current from the warmer water below and across the ice-ocean boundary layer. One way to estimate the magnitude of ε is to assume an approximate balance between the vertical turbulent heat fluxes that act to warm and cool, respectively, the boundary current^{34,49,51}. The former is often parameterised as a process of entrainment into the boundary current, whereby the entrainment rate is expressed as a function of the current speed and the boundary slope, while the latter is simply the heat flux that drives melting, defined above:

$$E_0 \Delta U \sin \theta (T_{*in} - T_*) \approx C_d^{1/2} \Gamma_{TS} U T_*$$

Both heat fluxes scale with the speed of the boundary current, in recognition of the fact that current shear drives the turbulent mixing. Using the above balance, the expression for ε becomes:

$$\varepsilon \approx \frac{2E_0 \sin \theta}{C_d^{1/2} \Gamma_{TS} + 2E_0 \sin \theta}$$

The neglect of heat advection within the boundary current in formulating the above balance of vertical turbulent fluxes is inappropriate when the inflow temperature is close to the surface freezing point and advection becomes critical in creating regions of basal freezing (where ε becomes negative). A pragmatic solution is to evaluate the thermal driving in the above theory relative to the surface freezing point and to define an offset at zero thermal driving, where the theory would give a melt rate of zero, from observations on ice shelves in cold oceanic environments, where a spatial average over regions of basal melting and freezing typically gives a small net melt rate²⁷.

Taking conventional values for physical constants ($g = 9.8 \text{ m s}^{-2}$, $f = -1.4 \times 10^{-4} \text{ s}^{-1}$, $L_i = 3.4 \times 10^5 \text{ J kg}^{-1}$, $c_i = 2.0 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$, $c = 4.0 \times 10^3 \text{ J kg}^{-1} \text{ K}^{-1}$, $\theta_s = 7.9 \times 10^{-4}$, $\theta_T = 3.9 \times 10^{-5} \text{ K}^{-1}$, $\lambda_1 = -5.7 \times 10^{-2} \text{ K}$) and parameters ($E_0 = 3.6 \times 10^{-2}$, $C_d^{1/2} \Gamma_{TS} = 5.9 \times 10^{-4}$), choosing appropriate mean values for Dotson Ice Shelf ($S_{in} = 34.5$, $T_{*i} = -15^\circ \text{C}$, $\sin \theta = 8 \times 10^{-3}$), and defining a zero thermal driving offset of 0.5 m yr^{-1} , yields the curve

plotted in Figure 4c. The shading around that line indicates the range of curves obtained when the parameters and input data for Dotson Ice Shelf are individually varied by $\pm 50\%$ (but $\pm 10\%$ in the case of S_{in} , because known variations are small, and $\pm 100\%$ for the zero thermal driving offset).

Data availability

The oceanographic data that support the findings of this study are available in the U.S. Antarctic Program Data Center (USAP-DC, <http://www.usap-dc.org/>), a member of the Interdisciplinary Earth Data Alliance (IEDA), with the identifier <https://doi.org/10.15784/601105>. Raw and processed data for individual cruises, along with details of the processing, can also be obtained upon reasonable request from the points of contact listed in Supplementary Table 1.

Code availability

The Matlab scripts used for the analyses described in this study, along with data files formatted for use with the software, can be obtained from the corresponding author upon reasonable request.

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Supplementary Information
for
West Antarctic Ice Sheet retreat in the Amundsen Sea driven by decadal
oceanic variability

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Supplementary Table 1 | Details of datasets used.

Year	Cruise Number	Ship	Contact	E-mail	Station numbers (west to east)	Dates	Instruments
2000	NBP0001	Nathaniel B Palmer	Jacobs	sjacobs@ldeo.columbia.edu	53, 54, 55	3-4 March 2000	CTD, DO, LADCP
2006	JR141	James Clark Ross	Jenkins / Shoosmith	ajen@bas.ac.uk	18, 17, 16, 15, 14, 13	30 Jan 2006	CTD
2007	NBP0702	Nathaniel B Palmer	Jacobs	sjacobs@ldeo.columbia.edu	64, 65, 66, 67, 68, 69	22-23 Feb 2007	CTD, DO, LADCP
2009	NBP0901	Nathaniel B Palmer	Jacobs	sjacobs@ldeo.columbia.edu	119, 120, 121, 122, 123, 124	4-5 Feb 2009	CTD, DO, LADCP
2011	NBP1005	Nathaniel B Palmer	Stammerjohn	sharon.stammerjohn@colorado.edu	29, 10, 11, 9	17 Dec 2010; 2 Jan 2011	CTD, DO
2012	ANA02C	Araon	Kim / Lee	twkim@kopri.re.kr	2701, 2801, 1901, 2001, 2101	16-18 Feb 2012	CTD, DO
2014	ANA04	Araon	Kim / Lee	twkim@kopri.re.kr	1026, 1025, 1024, 1023, 1022	8-9 Jan 2014	CTD, DO, LADCP
2016	ANA06	Araon	Kim / Lee	twkim@kopri.re.kr	221, 211, 201, 191, 181, 172	19-20 Jan 2016	CTD, DO, LADCP

Instruments are Conductivity-Temperature-Depth sensors (CTD), Dissolved Oxygen sensors (DO) and Lowered Acoustic Doppler Current Profilers (LADCP).

Supplementary Table 2 | End member properties used to calculate meltwater fraction for each year.

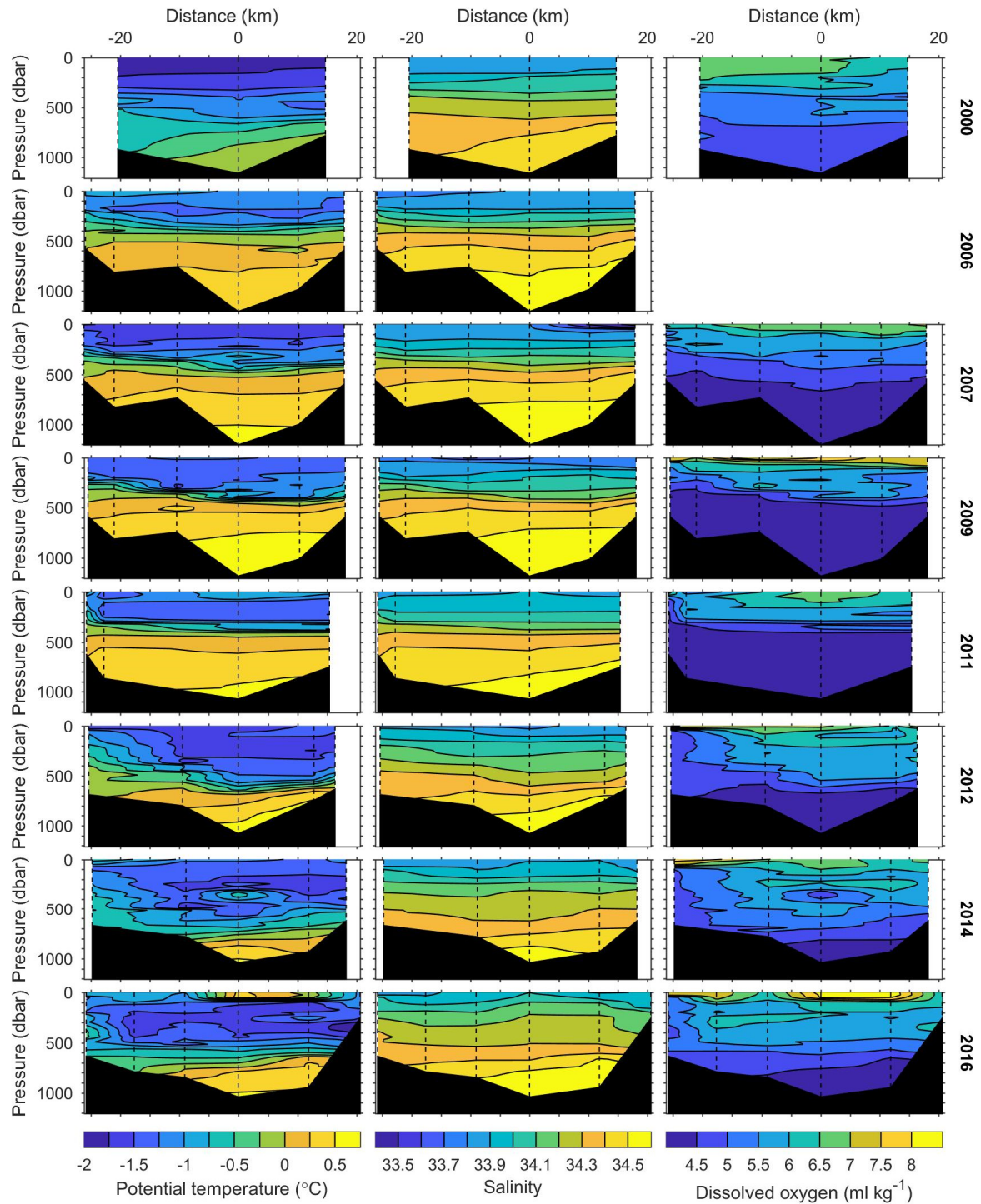
Year	Ice properties			mCDW properties			WW properties		
	T	S	DO	T	S	DO	T	S	DO
2000	-95	0	24.5	-0.3	34.43	4.75	$T_{fp}(S)$	34.22 – {0, 0.03, 0.06, 0.09}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2006	-95	0		0.75	34.63		$T_{fp}(S)$	34.07 – {0, 0.065, 0.13, 0.195}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2007	-95	0	24.5	0.8	34.62	4.36	$T_{fp}(S)$	34.18 – {0, 0.09, 0.18, 0.27}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2009	-95	0	24.5	1.0	34.65	4.225	$T_{fp}(S)$	34.1 – {0, 0.05, 0.1 0.15}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2011	-95	0	24.5	0.85	34.61	4.05	$T_{fp}(S)$	34.15 – {0, 0.06, 0.12, 0.18}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2012	-95	0	24.5	0.56	34.545	4.15	$T_{fp}(S)$	34.2 – {0, 0.035, 0.07, 0.105}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2014	-95	0	24.5	0.0	34.45	4.5	$T_{fp}(S)$	34.23 – {0, 0.02, 0.04, 0.06}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}
2016	-95	0	24.5	0.025	34.47	4.55	$T_{fp}(S)$	34.25 – {0, 0.015, 0.03, 0.045}	$DO_{sat}(S, T) -$ {0, 0.5, 1, 1.5, 2, 2.5, 3}

Properties are potential temperature (T in °C), practical salinity (S) and concentration of dissolved oxygen (DO in ml kg⁻¹). Ice properties are: an effective potential temperature that implicitly incorporates the heat loss associated with warming the ice by 15°C from its core temperature to the seawater freezing point, then melting it⁴²; a practical salinity of 0; a dissolved oxygen concentration that assumes the ice is formed from the compaction of cold firn at elevations of 1000–2000 m above sea level⁴⁷. Properties of mCDW come from observation, while those of WW are derived from an assumed salinity, with subscripts fp and sat indicating, respectively, the freezing point at atmospheric pressure as a function of salinity and the saturation concentration at atmospheric pressure as a function of salinity and potential temperature. For each year 28 realisations of WW properties are created by subtracting the values in braces from the maxima (shown graphically in Supplementary Figure 2).

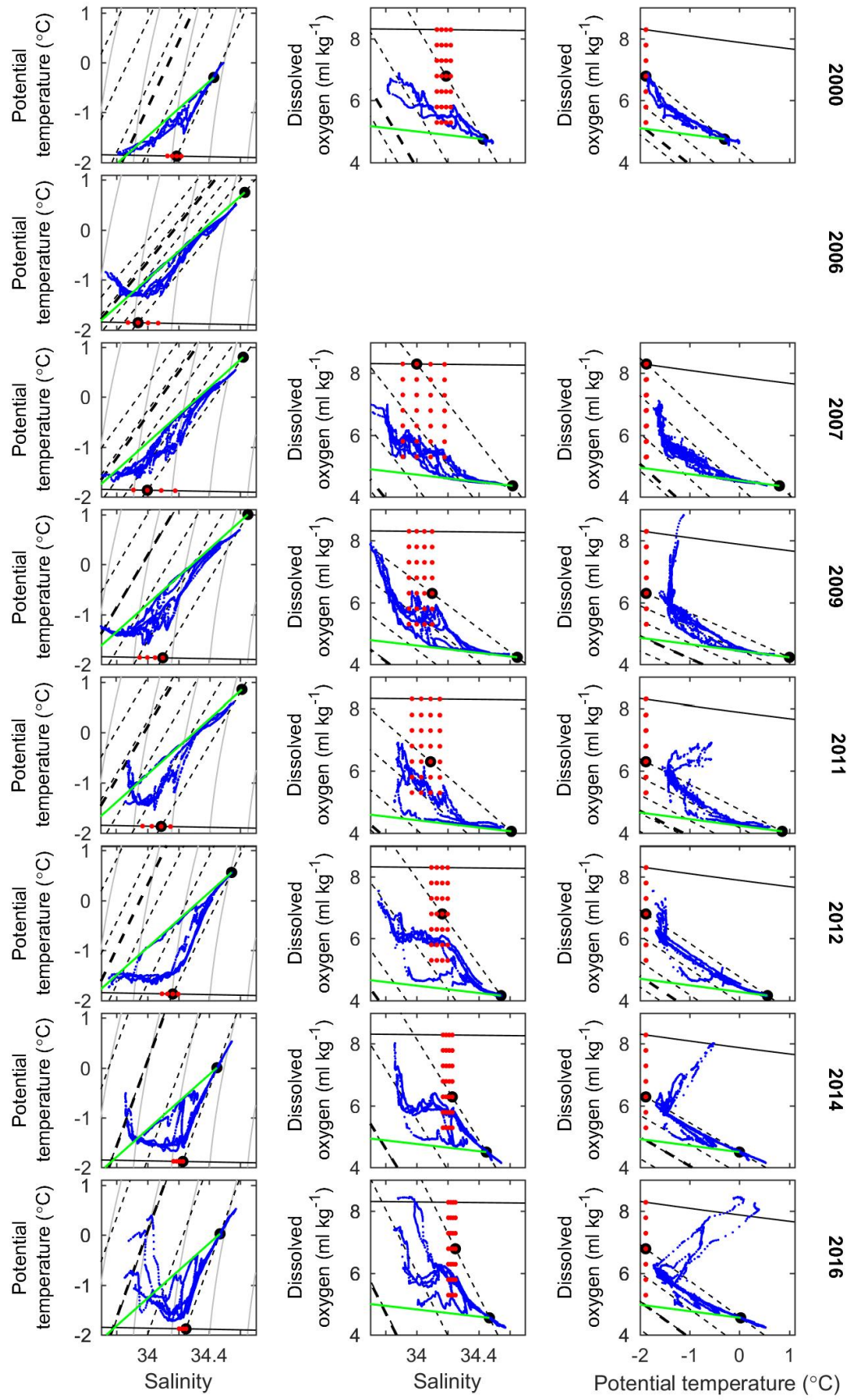
Supplementary Table 3 | Score-weighted mean and standard deviation for key diagnostics of the circulation across Dotson Ice Front in each year.

Year	2000	2006	2007	2009	2011	2012	2014	2016
Geostrophic solutions								
Number and type (F/T)	21 (F)	8 (F)	55 (F)	28 (T)	56 (T)	56 (F)	19 (T)	20 (F)
Mean inflow properties								
Transport (Sv)	1.17	0.39	0.61	0.71	0.46	0.64	0.83	0.79
STD (Sv)	0.28	0.05	0.19	0.37	0.07	0.06	0.12	0.24
Temperature (°C)	1.14	1.89	1.63	2.06	1.98	1.20	0.92	0.89
STD (°C)	0.09	0.19	0.19	0.22	0.12	0.27	0.04	0.13
Meltwater fraction (permille)	2.2	2.3	5.7	4.9	3.7	3.6	3.4	4.1
STD (permille)	1.0	0.9	1.9	1.8	1.1	2.3	1.2	1.5
Mean section properties								
Net transport (Gt yr ⁻¹)	31.0	65.6	31.0	88.2	51.5	20.9	23.1	20.5
STD (Gt yr ⁻¹)	2.5	5.1	10.8	23.4	7.0	3.3	6.9	10.8
Net melt transport (Gt yr ⁻¹)	20.8	45.9	33.8	97.2	56.2	19.7	30.6	18.5
STD (Gt yr ⁻¹)	4.4	10.1	16.7	30.8	9.8	5.1	7.8	10.9
Temperature (°C)	1.07	1.59	1.54	1.87	1.67	1.16	0.97	1.02
STD (°C)	0.09	0.10	0.20	0.20	0.15	0.14	0.02	0.03
LADCP solutions								
Number and type (F/T)	24 (F)	40 (T)		28 (T)	28 (T)			
Mean inflow properties								
Transport (Sv)	0.99	1.17		0.72	0.51			
STD (Sv)	0.46	0.49		0.18	0.09			
Temperature (°C)	1.15	1.70		2.20	1.14			
STD (°C)	0.09	0.34		0.13	0.06			
Meltwater fraction (permille)	1.7	4.9		4.2	2.0			
STD (permille)	1.2	1.6		1.0	1.2			
Mean section properties								
Net transport (Gt yr ⁻¹)	24.9	82.1		81.5	16.7			
STD (Gt yr ⁻¹)	6.1	21.9		12.7	3.3			
Net melt transport (Gt yr ⁻¹)	23.8	90.8		99.6	20.5			
STD (Gt yr ⁻¹)	8.1	37.9		17.3	2.9			
Temperature (°C)	1.12	1.64		1.85	1.02			
STD (°C)	0.10	0.25		0.18	0.08			

Number and type (F=full, T=truncated) of solutions with non-zero scores that contribute to the mean and standard deviation for each year. Geostrophic and LADCP solutions are shown separately above, but are combined in Supplementary Figure 6. Mean inflow properties are evaluated over all areas of the section where the inferred current is into the cavity, while mean section properties are evaluated over inflow and outflow regions. The near-surface layer where air-sea interactions affect properties (Supplementary Figure 3) was excluded from the calculations of circulation and property means. In cases where only a small region was excluded, calculations were repeated including the full water column from the sea surface down to the seabed, giving a maximum of 56 possible solutions for those sections. Temperatures are relative to the surface freezing point.

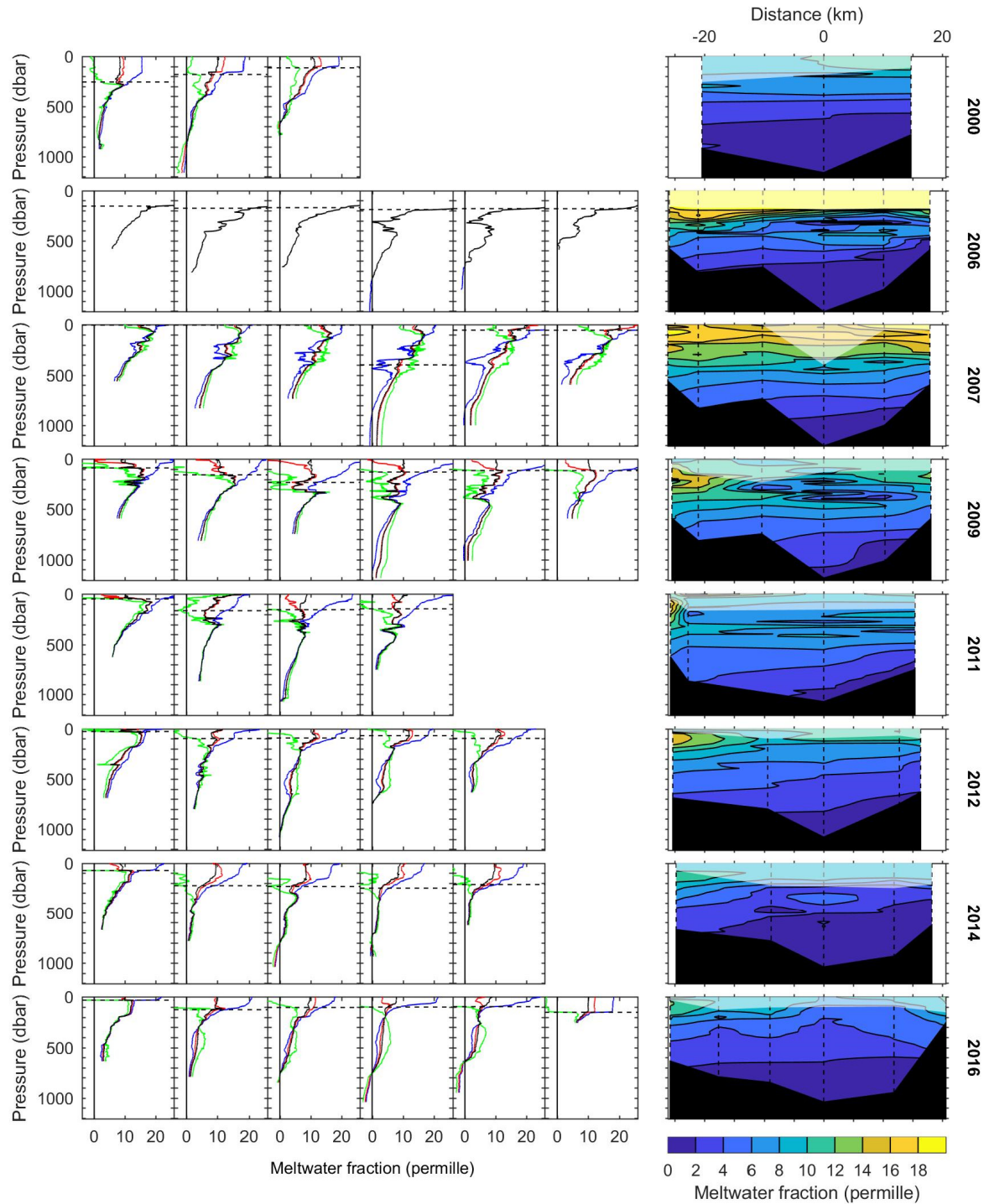


Supplementary Figure 1 | Cross-sections of potential temperature, salinity and dissolved oxygen concentration measured at Dotson Ice Front. Station locations are indicated by vertical dashed lines, with the year of observation on the right (Supplementary Table 1 has precise dates). Distance relative to the central trough station is positive eastward, and black shading shows the seabed, linearly interpolated between stations. Dissolved oxygen (right) was not measured in 2006.

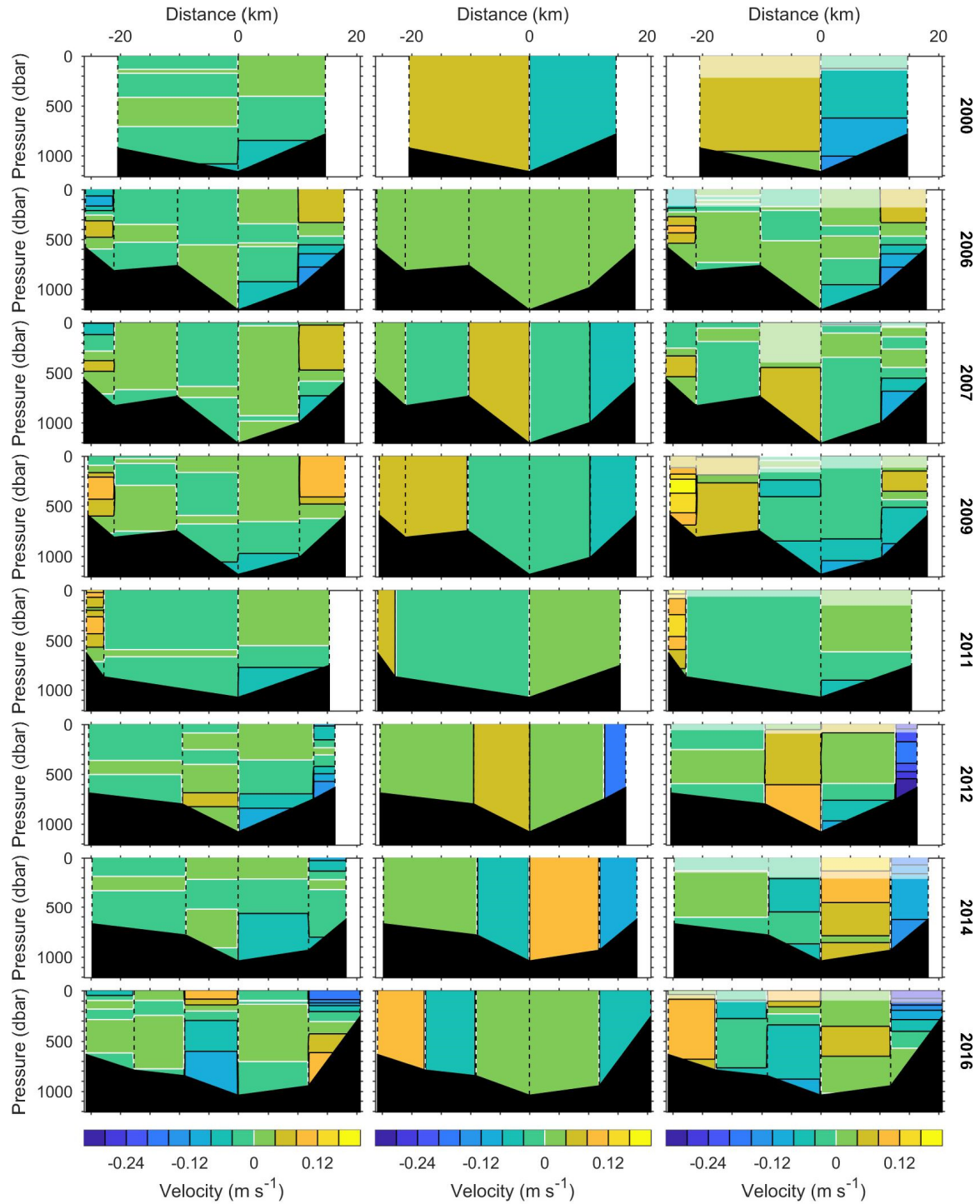


Supplementary Figure 2 | Scatter plots of potential temperature, salinity and dissolved oxygen concentration measured at Dotson Ice Front. Blue dots are 1 dbar averages from each vertical profile,

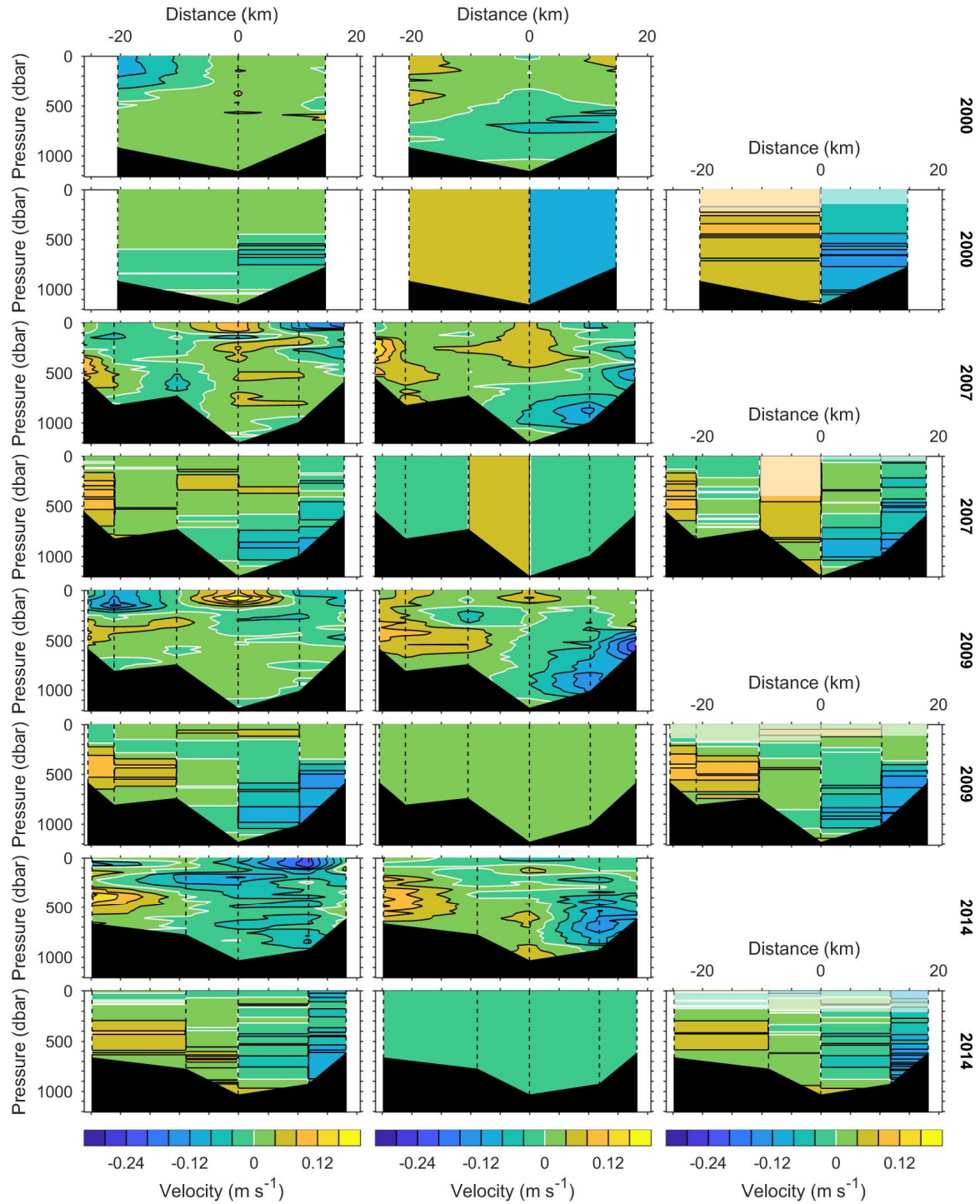
with the year of observation on the right. Dissolved oxygen was not measured in 2006. Solid black lines show the freezing point temperature (left-hand panels) and saturation concentration (centre and right-hand panels) at atmospheric pressure, while solid grey lines (left-hand panels) are contours of potential density (1027 to 1027.8 kg m^{-3} in 0.2 kg m^{-3} increments). Red dots indicate WW properties used in calculations of meltwater fraction and transport (Methods and Supplementary Table 2). Black circles are the mCDW properties (Supplementary Table 2) and the particular WW properties used to calculate contours of meltwater fraction (dashed, black lines, 0 to 40 permille in 10 permille increments) and results in Supplementary Figures 3, 4 and 5. Green lines connect mCDW properties and (assumed constant) ice properties⁴² (Supplementary Table 2). Bold, dashed, black lines are the theoretical upper bounds of meltwater fraction (intersection of green and freezing point lines)⁴².



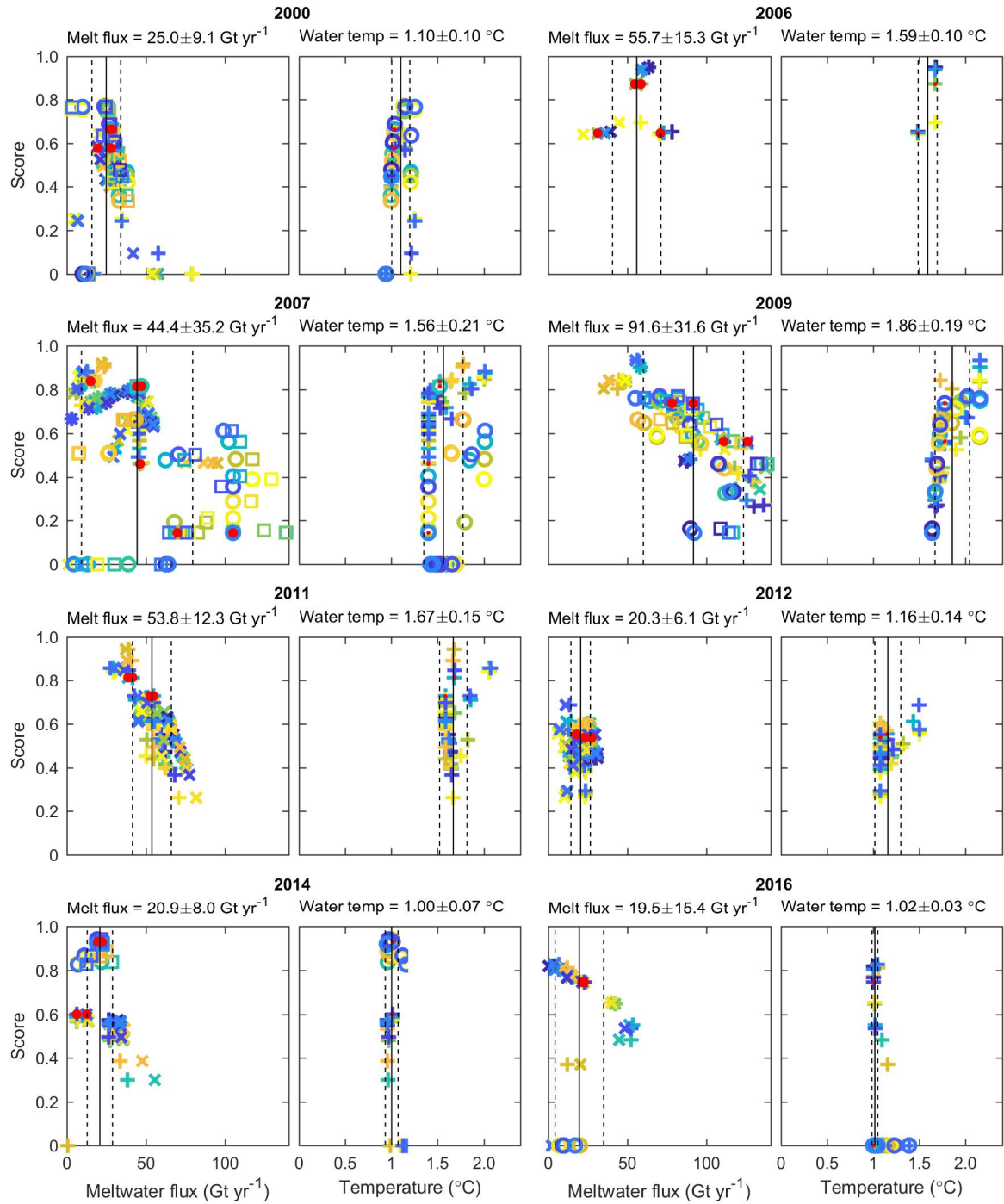
Supplementary Figure 3 | Individual station profiles and resulting cross-sections of meltwater fraction from observations at Dotson Ice Front. Three estimates of meltwater fraction (Methods and Supplementary Figure 2) are based on potential temperature and salinity (blue profiles), dissolved oxygen and salinity (red profiles) and dissolved oxygen and potential temperature (green profiles). Black profiles show the mean of those estimates (only one is possible in 2006), used to produce the contoured cross-sections (right). Dashed horizontal lines on profile plots and lighter shading on cross-sections indicate where meltwater fractions are disregarded because air-sea interactions affect near-surface properties. Horizontal distance axis origin and direction as in prior figure; individual station profiles also from west (left) to east (right).



Supplementary Figure 4 | Geostrophic currents perpendicular to Dotson Ice Front. Geostrophic currents between each station pair (left-hand panels) calculated using the slope of the isobars, determined from temperature and salinity profiles (Supplementary Figure 1), and an assumption of zero velocity at the lowest level in the water column with a significant meltwater fraction (Methods, Supplementary Figure 3). Depth-independent velocity adjustments (centre panels) are the minimum additions required for the resulting geostrophic circulation (right-hand panels) to satisfy integral constraints on the net transport of heat, salt and dissolved oxygen (Methods, particular solution plotted was obtained using WW properties indicated in Supplementary Figure 2). Positive values indicate flow away from the ice (northward), with lighter shading (right-hand panels) indicating the near-surface layer excluded from the calculations because of the impact of air-sea interaction.



Supplementary Figure 5 | Directly-measured currents perpendicular to Dotson Ice Front in 2000, 2007, 2009 and 2014. In each summer, zonal (left) and meridional (right) Lowered Acoustic Doppler Current Profiler components are shown in the upper two panels. The lower three panels show mean currents perpendicular to the ice front between each station pair (left) and minimum depth-independent velocity adjustments (centre) required for the resulting circulation (right) to satisfy integral constraints on the net transport of heat, salt and dissolved oxygen (Methods, particular solution plotted was obtained using WW properties indicated in Supplementary Figure 2). Positive values and lighter shading (right) as in previous figure. LADCP data in 2016 sampled a strong transient, depth-independent flow at one station, preventing any solutions with scores above zero (Methods and Supplementary Figure 6).



Supplementary Figure 6 | Meltwater flux and mean temperature above the surface freezing point across Dotson Ice Front. Results obtained for each choice of WW properties (Supplementary Table 2 and Supplementary Figure 2) are plotted (colour-coded) against the score for that solution (Methods). Solutions derived from geostrophic currents are indicated by crosses (+ and x), and those from LADCP data by circles and squares. On plots of meltwater flux, crosses (+) and circles indicate the difference between inflow and outflow, and crosses (x) and squares the net meltwater flux (Methods). Red dots denote solutions obtained with WW properties highlighted in Supplementary Figure 2 and used in Supplementary Figures 3, 4 and 5. Score-weighted means and standard deviations are indicated by vertical solid and dashed lines, respectively, and are noted numerically above each plot. Mean temperatures are averaged over the relevant section area included in the constrained budget calculation. Final score-weighted results are used in Figure 4 of the main text.